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Scaling and Effects of Surface Inhomogeneities

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Diffusion Modeling with Convective Scaling and
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16. ABSTRACT The development of convective scaling is reviewed, along with results from its application to turbulence and diffusion data. Further analyses of a wide variety of diffusion data for σ_y and crosswind-integrated concentration are presented in nondimensional form with convective scaling, with remarkably consistent results. Preliminary data from the CONDORS convective diffusion experiment, performed at the Boulder Atmospheric Observatory in Colorado in 1982 and 1983, are included. For elevated sources, an enhancement of peak surface concentrations over those predicted by conventional Gaussian modeling is found; the effect increases as the ratio of source height to mixing depth increases. Later sections of this paper assess the effects of surface inhomogeneities on diffusion in convective conditions. Using the ratio w^*/U , which characterizes initial rate of spread, effects typical of large urban areas and forests on diffusion rates are estimated (w^* is the convective scaling velocity, and U is the mean wind speed). Estimated fetches over changed surface conditions for nearly full readjustment are roughly $(U/u^*)^2 z_i$ for U and $2(U/w^*)z_i$ for turbulence velocities (u^* is friction velocity and z_i is the mixing depth). Finally, examples of heat flux inhomogeneities causing stationary convective features to develop are presented. It is suggested that the time averaged downdraft speeds associated with these features are about $0.4w^*$; such downdrafts could significantly increase surface concentrations from elevated sources.		
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Abstract

The development of convective scaling is reviewed, including some results of its application to turbulence and diffusion data. When this scaling applies, the most significant length scale is the mixing depth, z_i ; velocities scale to $w^* = (H^* z_i)^{1/3}$, where H^* is the flux of buoyant acceleration due to heat and water vapor fluxes at the surface; the time scale for diffusion through the convective boundary layer is z_i/w^* . New analyses of a wide variety of diffusion data for σ_y and crosswind-integrated concentration are presented in nondimensional form using convective scaling. These include some preliminary data from the CONDORS convective diffusion experiment, performed at the Boulder Atmospheric Observatory in Colorado in 1982 and 1983. For elevated sources, an enhancement of peak surface concentrations over that predicted by conventional Gaussian modeling is noted; the effect increases as the ratio of source height to z_i increases.

The effects of surface inhomogeneities on diffusion in convective conditions are considered. Using the ratio w^*/U to characterize the initial rate of spread, effects typical of large urban areas and forests on diffusion rates are estimated; these effects include sometimes doubled H^* , with 15 to 30% increases in z_i and in U^{-1} , where U is the mean wind speed in the mixing layer. Small cities do not significantly change z_i and U . Estimated fetches over changed surface conditions required for nearly full readjustment are roughly $(U/u^*)^2 z_i$ for U , where u^* is friction velocity, and $2(U/w^*) z_i$ for turbulence velocities. Finally, examples of H^* inhomogeneities causing stationary convective features to develop are

discussed. The time averaged downdraft speeds associated with these are about $0.4w^*$, calculated using H^* differences; such downdrafts could significantly increase surface concentrations from elevated sources.

1. Introduction

For the past several decades, two methods have been widely used for classifying atmospheric stability to order diffusion data and make diffusion predictions. Pasquill (1961) proposed wind speed and insolation or nighttime cloudiness as classifiers, with wind speed having the stronger effect. Briggs (1982a) showed that wind speed cubed divided by insolation strongly correlates with the daytime Obukhov length, and this length has been shown by a number of investigators to order vertical diffusion near the ground most effectively (see Briggs, 1982b). The "Brookhaven method" (Smith, 1973) uses a single stability index, the variance of horizontal wind direction, σ_a . It is not surprising that σ_a correlates well with lateral dispersion, σ_y , because the "waved garden hose" analogy implies that $\sigma_y \approx \sigma_a x$ near the source (x = distance downwind of the source). More surprising is the observation that, in the daytime and near the source, σ_a also correlates well with vertical dispersion, whereas at nighttime this correlation is nil (Cramer *et. al.*, 1958). A third classifying method, using temperature gradient alone, was shown to be essentially uncorrelated with daytime diffusion rates (Weber, *et al.* 1977; Briggs and McDonald, 1978). It correlates strongly with surface heat flux, but has little correlation with daytime wind speed, which is the stronger factor affecting diffusion.

An entirely new approach to predicting daytime diffusion has grown out of scaling ideas proposed by Deardorff (1970a), who was inspired by Busse's (1970) scaling of convective turbulent motions generated by a constant heat flux between two horizontal plates. Deardorff suggested the following for atmospheric boundary layers dominated by convectively

driven turbulence: (1) large eddy motions in a convectively mixed layer scale to the mixing depth, z_i , (2) turbulent velocities are proportional to a convective velocity scale, $w^* = (H^* z_i)^{1/3}$, and (3) temperature fluctuations from the mean are proportional to $\theta^* = \overline{w'T'}/w^*$. Here, H^* is $(g/c_p \overline{\rho} T)$ times the surface sensible heat flux, which is equivalent to $(g/\overline{T}) \overline{w'T'}$ (g is gravitational acceleration, c_p is the specific heat capacity of air, ρ is the density of air, T is the absolute temperature of air, and w is the local vertical velocity; bars denote time means, and primes denote deviations from the mean). This is simply the vertical flux of buoyant accelerations produced by small temperature fluctuations; alternatively, it can be understood as flux of buoyant force divided by the density of the medium. (A simple way to derive w^* and θ^* is to consider the rise of a single thermal with a vertical velocity $\overline{w}$ that carries an acceleration flux proportional to H^* . The buoyant acceleration is then $\propto H^*/\overline{w}$, and assuming that the kinetic energy is proportional to the potential energy released as the thermal traverses z_i , we have $\overline{w}^2 \propto (H^*/\overline{w}) z_i$. This gives $\overline{w} \propto w^*$, and the temperature excess of the thermal is $(T/g)(H^*/\overline{w}) \propto \theta^*$.)

Deardorff (1970b) demonstrated that some turbulence variances measured by aircraft agree well with those predicted by his numerical model of the boundary layer for sufficiently convective conditions when scaled by w^* and θ^* . That is, the dimensionless coefficients are nearly identical. The same scaling ideas were applied to an extensive set of boundary-layer turbulence measurements made in Minnesota in 1973, with good results (Kaimal *et al.*, 1976). The most striking improvement was in the ordering of the horizontal turbulence velocity spectra, which previously had eluded any ordering in unstable conditions when analyzed

in terms of surface layer similarity theory (Kaimal et al., 1972). The difference between convective scaling and Monin-Obukhov surface layer similarity scaling amounts to the use of z_i in place of the Obukhov length, $L = -u_*^3/(kH_*)$, where u_* is the friction velocity and k is the von Karman constant. Note the parallel relationships in the two velocity scales: $u_* \propto |LH_*|^{1/3}$ and $w_* = (z_i H_*)^{1/3}$. Both scalings use the surface buoyancy flux.

The Minnesota results were reinforced by a second experiment made in England over very different topography (Caughey and Palmer, 1979). Further proof of the effectiveness of convective scaling came from laboratory experiments in a water tank, 1.2 m square, with a uniformly heated bottom (Willis and Deardorff, 1974). The buoyancy flux was appropriately redefined using the coefficient of thermal expansion for water in place of T^{-1} . The width-to- z_i aspect ratios of 2 or 4 were a restrictive factor, as were the relatively low Reynolds numbers. Nonetheless, the velocity and temperature fluctuations generally agreed well with aircraft turbulence data, after they were scaled according to w_* , θ_* , and dimensionless height, z/z_i .

2. Convective scaling applied to diffusion

Because turbulent motions are responsible for all significant diffusion in the atmosphere, the success of convective scaling in ordering turbulence data offers much hope that it will lead to improved methods for predicting diffusion in convective conditions. Seeing this possibility, Willis and Deardorff (1976, 1978, 1981) used their modeling tank to investigate lateral and vertical diffusion from passive line sources. They photographed

neutral-buoyancy oil droplets along the axial direction for three release heights, $z_s/z_i = 0.067, 0.24, \text{ and } 0.49$, where z_s is the source height. Their results for lateral spread were more or less as expected, with nearly Gaussian distributions in the ensemble mean. At small distances, the lateral variance, σ_y , was approximately proportional to w^*t , where t is the time from release. This is consistent with the statistical theory prediction for small t that $\sigma_y = \sigma_v t$ and one of the most useful results from the Minnesota experiment: in moderate to strong convection, it was observed that the lateral velocity variance $\sigma_v \approx 0.6w^*$ throughout most of the mixing depth, even near the ground (Caughey and Palmer, 1979). At large dimensionless time, w^*t/z_i , Willis and Deardorff observed a tendency towards $\sigma_y \propto (w^*z_i t)^{1/2}$, which is consistent with a kinematic viscosity $\propto w^*z_i$.

At small w^*t/z_i , Willis and Deardorff's (1976) results for vertical diffusion from a near-surface source are not surprising. The plume was nearly half-Gaussian, with the vertical diffusion coefficient, σ_z , approximately proportional to $H^{1/2}t^{3/2}$. This is consistent with "free" convective scaling, in which z is the only significant scaling length (a free convective regime exists if z is below $0.1 z_i$ and above $-L$; thus it exists only if $z_i > -10L$). This result can explain the good correlation of σ_z with σ_a for a ground source in convective conditions. The above proportionality for σ_z can be expressed in terms of σ_a by substituting $H^* = w^{*3}/z_i$ and $w^* \propto \sigma_v \approx \sigma_a \bar{u}$, where $\bar{u}$ is the mean wind speed at the height of σ_a measurement. Since wind shear is not large in convective conditions, $\bar{u}t \approx x$, and, approximately, $\sigma_z \propto (\sigma_a x)^{3/2} z_i^{-1/2}$; $z_i^{1/2}$ does not vary nearly as much as $\sigma_a^{3/2}$, hence the good correlation of σ_z with σ_a , as found by Cramer (1958).

Just beyond $w^*t/z_i = 0.5$, diffusion in the tank departed from its expected behavior. Surface concentrations for the near-surface source began to drop rapidly, as the maximum concentrations lifted off the surface, to be found later at about $0.75z_i$. At this point, the Gaussian-plume assumption fails, and the eddy diffusivity derived from the time and space derivatives of the concentration field behaves strangely, taking on negative values for part of the time (Deardorff and Willis, 1975). The apparent physical cause is that surface fluid is drawn into thermals and picked up while being replaced by relatively "clean" flow from downdrafts. The rapid fall in ground concentration at $t > 0.5z_i/w^*$ explains the explosive apparent growth of σ_z when the Gaussian plume assumption is applied to surface measurements from surface sources. Such was the case for the early analyses of the Prairie Grass field experiment, which strongly influenced Pasquill's (1961) σ_z curves for "A" and "B" stability categories.

In the meantime, Lamb began using results from a later version of the numerical model of Deardorff (1972) to predict diffusion in the convective boundary layer. Since this "large eddy" model predicts actual eddy motions, plus eddy diffusivity attributed to smaller than grid cell sized motions, Lamb "released" thousands of particles into this numerical field and traced their motions to derive concentration fields. He did this for release heights $z/z_i = 0.026, 0.26, 0.52, \text{ and } 0.74$, and then nondimensionized the results using convective scaling so that they could be compared to laboratory and field experiments (Lamb, 1979). His results for the near-surface release were remarkably similar to those of Willis and Deardorff's tank experiments, as discussed above.

For the more elevated releases, Lamb's results took another unexpected turn. Rather than remaining near the release height, the maximum concentrations in the ensemble mean descended, at a rate $\approx 0.5w^*$; this resulted in larger surface concentrations than given by Gaussian models with the plume centerline remaining at source height. After impacting the surface, the concentration patterns behaved much like that from the low level release, with the maximum lifting off the surface at larger values of w^*t/z_i . These unanticipated concentration patterns led to the elevated source laboratory tank experiments by Willis and Deardorff (1978, 1981). Once again, the essential patterns predicted by the numerical model were confirmed in the laboratory experiments.

Because the numerical model simulated actual turbulent motions for the larger eddies, Lamb was able to investigate the cause of the descending maximum from elevated sources. He concluded that it occurs simply because downdrafts, though slower than updrafts, occupy more horizontal area than updrafts (thermals). Therefore, the larger part of the source material is emitted into downdrafts. The Lagrangian time scale is relatively large; that is, particle trajectories tend to continue as they start out, until reflected by the ground or z_i . Turbulence is also less in downdrafts, so material traveling within a downdraft tends to diffuse less within it than material released into an updraft.

In Lamb's numerical modeling, there was a crosswind, and the tracers were started continuously from a point, rather than instantaneously from a line. For this case, t is replaced by x/U in the scaling analysis, where U is the mean wind speed in the mixing layer (fortunately, wind shear is normally quite small through most of a convective mixing layer, due to

the vigorous vertical mixing of horizontal momentum). Dimensionless time is replaced with dimensionless x/U , defined by $X = (w^*/z_i)x/U$. Continuous point-source diffusion differs from line source diffusion in that longitudinal diffusion could smear the concentration distribution along the x axis. This is especially significant when U/w^* is less than 1.5 (Deardorff and Willis, 1975).

A few investigators have looked at existing field data in terms of convective scaling. However, of the early diffusion experiments, only the Prairie Grass program (Barad, 1958) included vertical soundings that allowed z_i to be determined for each run; z_i is the single most important parameter for convective scaling. An estimate of H^* can be wrong by a factor of 2 without dire consequences; this would change w^* and X determinations by a factor of only 1.26. The detailed profiles of wind speed and temperature in the lowest 16 m permitted H^* to be estimated with some confidence, using established flux-profile relationships. Briggs and McDonald (1978) first showed that the crosswind-integrated concentrations observed on the Prairie Grass surface arcs were consistent with surface concentrations observed in the tank experiment when both were nondimensionalized through convective scaling. Nieuwstadt (1980) took this type of analysis further, showing that the data fit the free convection prediction over most of their range in X , with $U \int \chi dy \propto QH^{*-1/2}(x/U)^{-3/2}$. He further showed that the Prairie Grass σ_y values closely paralleled the tank and numerical model results when convectively scaled, although the field values averaged a factor of 1.4 larger. Nieuwstadt pointed out that horizontal motions in the tank are inhibited by the tank walls, which may diminish σ_y values.

Several modelers have adapted convective scaling ideas and results to plumes from power plants. Weil (1979) took the expressions for σ_y suggested by Lamb (1979) and combined them with plume growth expected from buoyant plume rise, with plume rise and plume diameter both proportional to $F^{1/3}U^{-1}x^{2/3}$ (F is the source buoyancy flux parameter, the point-source equivalent of H^*). Weil's expression for σ_y agreed well with lidar and mobile van (correlation spectrometer) measurements made at the Morgantown power plant, in Maryland. Weil's expression introduced the dimensionless ratio $F/(Uw^*z_i)$. In a later series of tank experiments, this time with buoyant plumes, Willis and Deardorff (1983) found this ratio to be an important measure of source buoyancy effects in the convective layer. More recently, Weil and Brower (1982) developed a more elaborate model for tall-stack plume dispersion that incorporates convective scaling parameters in both σ_y and σ_z , as well as in the plume rise expressions; they demonstrated its clear superiority over a standard modeling approach in predicting ground concentrations observed at three power plants. Venkatram (1980) also introduced a model that used convective scaling parameters to predict σ_y and σ_z and included buoyant plume rise, plume growth due to plume rise, and impingement on the ground caused by convective downdraft velocities proportional to w^* . He compared the predictions of this model with ground concentrations measured at two power plants. More recently, Venkatram and Vet (1981) tested a simplified version of this model against the same power plant data and also against data from a smelter with a 380-m stack, with good results. This version neglects plume growth due to plume rise and approximates σ_y with $0.45(w^*/U)x$.

3. New and old field data for convective scaling analysis

Although buoyant plume dispersion is probably the most important potential application of convective scaling models, the modeling-tank and numerical results for passive sources need to be tested in the real atmosphere. This is especially true for elevated releases, as well as for surface releases at distances greater than 0.8 km. An attempt to partially fill this need was begun last fall at the Boulder Atmospheric Observatory (BAO) of NOAA's Wave Propagation Laboratory. To map plume concentrations in three dimensions through the height z_i and to several $(U/w^*)z_i$ downwind, two different remote sensor techniques were tried, with hopes of eventually mapping an elevated and a surface release at the same time, independently. The 1982 releases were made side by side, to compare the two techniques directly. One release was a highly visible oil fog, measured by lidar; the other was metallic-coated chaff, detected by X-band radar. Details of instrument siting and signal processing are given by Moninger et. al., (1983). An outstanding advantage of the BAO site was the well-instrumented 300-m meteorological tower, with a carriage that permitted a release height to be selected just before a run.

From three two-hour releases made in 1982, five periods with relatively constant wind speeds and directions were chosen for processing. Only the chaff results are complete at the time of this writing (chaff and oil fog results for one period are compared in Moninger et. al., 1983). Radar returns are not affected by oil fog, but they are affected by any objects with metallic content. Therefore, most of the signal returned from the lowest 50 m had to be ignored because of "ground clutter." The results given here for "surface" concentrations are actually for the 50- to 100-m

layer. This loss of information was especially significant for period 2, which had a mixing depth of only 500 m. A second shortcoming was that the metallic chaff's settling velocity, about 0.3 m/sec, was not negligible compared to convective turbulence vertical velocities, of the order of 1 m/s. This effect has not yet been completely assessed. More extensive measurements were made at the BAO in 1983, and will provide better information when processing and analysis are completed.

To demonstrate the universality of convective-scaling results, the BAO chaff data will be nondimensionalized and plotted along with data from the literature; these include Prairie Grass surface SO_2 releases (Barad, 1958), elevated SF_6 releases made at Cabauw, in the Netherlands (Agterberg *et. al.*, 1983), and plume measurements at the Morgantown power plant (Weil, 1979). Also to be shown are curves derived from the tank experiments (Willis and Deardorff, 1976, 1978, 1981) and from the numerical modeling experiments (Lamb, 1979). Table I gives the parameters measured during the BAO periods. Mixing depth was determined from the vertical distribution of chaff at the longer distances. At each distance, the second highest value of $\int \chi dy$ in 50-m layers was found, and then the heights at 1/2 and 1/6 this value were interpolated; these two heights were used to linearly extrapolate a height for $\chi = 0$. For consistency with Willis and Deardorff's reckonings of z_i , z_i was taken to be 0.92 times this zero-projection height. The chaff settling velocity probably reduces concentration values in the upper part of the mixed layer, but should not have a large effect on the "zero-projection" height. H^* was determined from the lowest two levels of $\overline{w'T}$ measured on the BAO tower, except for 16 Sept., for which these data were missing. On the other days, there was

a good correlation between $\overline{w'T'}$ and $S^{3/2}$ (where S is the solar radiation); this correlation was used to estimate $\overline{w'T'}$ for 16 Sept. The median of the sonic anemometer measurements at 150, 200, 250, and 300 m was used for U . Although the integrated chaff return, $\iint \chi dydz$, showed no trend with distance after the first kilometer, it was not considered a conservative tracer. Instead, Q was replaced by $U \iint \chi dydz$ in the nondimensionalizations. Note that period 6 had an unusually small U/w^* ; in this case, longitudinal diffusion could have been important. In addition, some artificial longitudinal diffusion appears in the present analysis, because 500-m running averages were used to smooth out the bumpy character of the measured chaff field. The duration of period 6 was inadequate in terms of the time scale for passing of large eddies, about $1.5 z_i/U$; the durations of periods 4 and 5 were only marginally adequate.

All the BAO runs were quite convective, but for the other data sets, weakly convective runs were screened out of this analysis using the ratio U/w^* , which is approximately proportional to u^*/w^* and $|L/z_i|^{1/3}$. For the Prairie Grass data, wind measurements were available only from a 16-m tower. The measured or interpolated value at $z = 8$ m was used for U , and runs with $Ug/w^* > 5$ were excluded from the figures. The SO_2 releases were only 10 min long, which, in terms of eddy passage time is inadequate for good ensemble averaging. Consequently, runs were relatively susceptible to effects from individual eddies, and scatter was increased. Also, because the arcs extended only to $x = 800$ m, these runs did not extend far into distance scaled by $U(z_i/w^*)$.

The Cabauw experiments consisted of SF_6 releases made at 80 and 200 m. I chose to reevaluate z_i for some of the periods, relying mostly on the

rawinsonde potential temperature profiles, rather than using the evaluations offered by van Duuren and Nieuwstadt (1980). Because $\overline{w'T}$ was available for only a few of the runs, I used these runs to calibrate the relationship $H^* \propto R^{3/2}$, where R is the net radiation, which was available for all runs (the same power law was found approximately to fit the BAO heat flux and solar radiation measurements). Wind speeds measured by cup anemometers at 160 and 200 m on the Cabauw meteorological tower were used for U . Although runs were made on 14 days throughout the year, most were in neutral or only slightly convective conditions. Only the 3 days with $U/w^* < 6.7$ are used. There were five 30 min sampling periods, with surface samplers on only a single arc ranging from 3150 to 4300 m from the source.

The Morgantown data consisted of plume measurements made by stationary lidar and by a mobile van, which was equipped with a correlation spectrometer to measure the overhead burden of SO_2 . The van made about six traverses under the plume during each 1 h averaging period. The plant has two 213-m stacks spaced 76 m apart, each emitting about 80 MW of heat. To eliminate data for which plume buoyancy had a strong effect, I used the ratio $(F/x)^{1/3}/w^*$ as a screening index (during the rise stage, the plume vertical velocity, as well as its rate of growth due to entrainment, is proportional to $(F/x)^{1/3}$). All run parameters for Morgantown are listed by Weil (1979). Table 3 summarizes the ranges of parameters for all four sets of field data plotted in Figs. 1-4.

4. The universality of convectively scaled diffusion

This section attempts to show that convective scaling of diffusion observations provides consistent results. "Consistency" and "universality" are relative terms, of course, but the results should be judged considering the diversity of the types of observations being employed.

Figure 1 shows σ_y nondimensionalized by z_i and plotted against X . The tank-experiment curves and the numerical model curves are in good agreement. Run 1 from Prairie Grass, which is flagged, stands apart from the rest of the data. The lateral concentration distribution shows a double mode for this case, which suggests a wind shift during the run. The rest of the Prairie Grass points, which do not extend beyond $X = 0.4$, could be approximated by the small x asymptote $\sigma_y \approx \sigma_y t$ with $\sigma_y = 0.6 w^*$, namely, $\sigma_y = 0.6 w^* x / U$. The Morgantown power plant data points, which extend to 10 times as large X , cluster around the tank and numerical-model elevated-source curves. Only points with $F/(xw^*3) < 0.2$ were used, so that the enhancement of σ_y by buoyant rise should not exceed about 20%. These points are for the vertically integrated plume; most of the mobile van measurements showed similar σ_y values at the surface, but three of the surface measurements were about twice the σ_y of the vertically integrated plume.

Of the five periods at the BAO, the two surface releases agree well with the tank-experiment numerical-model curves, as does period 4 (open triangles), with $z_s/z_i = 0.17$. The σ_y 's of periods 2 and 3 (open circles and squares) run about twice as high, closer to the small- x asymptote. Both these periods are from the same 2-h run, made on a day with considerable wind direction shear in the mixed layer--consistently 13° from $z = 100$ m

to 300 m on the tower, with another 40 to 50° direction shear to the top of the mixed layer. Wind direction shear may enhance σ_y in conditions of weak to moderate convection. A measure of convective strength, $|z_i/L|$, was estimated to be 20 to 30 for these periods, in contrast to 180 and larger for the other three periods. Only the σ_y 's of the vertically integrated plumes are shown here; σ_y 's in the lowest 200 m are 20 to 40% smaller for periods 2 and 3, and are about the same as for the vertically integrated plumes in the other three periods.

Figure 2 shows crosswind integrated concentrations at or near the surface, nondimensionalized by z_i , U , and Q ; this was designated C_y , and plotted against X . For elevated sources, only C_y values at $x > 2.9 z_s(U/w^*)$ are plotted in this figure. This is downwind of the peak surface concentrations, where the elevated source plumes behave more like surface source plumes (Fig. 3 shows C_y for all x for the elevated sources). For the elevated-source case, the tank experiment and numerical modeling curves agree well, but for the near-surface releases, the tank curve dips deeper into subunity values of C_y . Most of the Prairie Grass data approximate the free convective law, as was shown by Nieuwstadt (1980); the mixing depth cancels out of the expression $C_y \propto X^{-3/2}$ when w^* is replaced by its definition. Only the four Morgantown points with $F/(xw^{*3}) \leq 0.032$ and surface scans of SO_2 are included (I assumed that $\int \chi dy = (2\pi)^{1/2} \sigma_y$ times the axial concentration). These points show an approach to unity C_y , which corresponds to uniform vertical mixing through z_i . The remaining Morgantown points, with $F/(xw^{*3}) = 0.056$ to 0.095 and $X = 0.35$ to 0.56 , had values of C_y ranging from 0.34 to as low as 0.045 , one to two orders of magnitude smaller than values for near-surface, passive sources. This

was likely due to buoyant plume rise and resistance to downward mixing. The two Cabauw points fall close to the elevated source curves, as do the points from the BAO chaff releases for the three elevated-source periods. The C_y values of the two surface source periods reach a smaller minimum value, the same trend indicated by the tank-experiment curves, but the curve for period 6 dips even lower, to $C_y \approx 0.4$. The effect of the fall velocity of the chaff is not yet known, but it seems likely that it should cause increased concentrations near the ground. It should be remembered that period 6 had an unusually small value of U/w^* , and its dimensionless duration was inadequate for good averaging.

A new convective scaling representation is shown in Fig. 3, with the near-surface crosswind integrated concentration nondimensionlized by z_s , U , and Q ; this was designated C_y' and plotted against $X' = (x/U)w^*/z_s$. This plot is convenient for elevated, passive releases, because C_y' peaks at about $X' = 2$, regardless of source height. A peak could occur at $X' = 2$ if a plume diffused in Gaussian fashion from its source level with $\sigma_z = 0.5w^*x/U$, or if the maximum concentration descended a distance $0.5w^*x/U$ with no diffusion at all. The results indicate a combination of these scenarios. For a Gaussian plume centered at $z = z_s$, C_y' would attain a maximum value of $(2/\pi e)^{1/2} = 0.48$, regardless of the rate of spread. The tank-experiment curve for the near-surface case peaks near this value, but the curves for larger $z_s/z_i = Z_s$ show a monotonic increase in the peak C_y' . This is also seen in the numerical-modeling curves and in the BAO chaff data. At small X' , the C_y' values for the chaff appear to climb more rapidly than the tank experiment curves, but this may result from the height at which measurements were taken, which averaged

75 m; this is about 1/3 the source height for periods 2 and 3. The Cabauw data for runs 1a and 1b and for runs 9a and 9b were combined into 1 hr averages, to reduce scatter. All of the data in Fig. 3 show good ordering with Z_s . Note that as uniform vertical mixing is approached, C_y' approaches Z_s .

The effect of Z_s on the peak value of C_y' is highlighted in Fig. 4. Period 4 of the BAO data was omitted, because chaff measurements were not made close enough to the source to identify the peak in C_y' , and the nearest radar returns were questionable because of ground clutter, possibly caused by the tower's guy wires. The one Cabauw point, for run 1, is presumed to be almost at the peak value, because $X' = 2.55$, but with only one sampling arc, a peak value cannot be identified. There is no significant trend in the X' of maximum C_y' with Z_s . Except for the numerical model's value for $Z_s = 0.74$, the maximum C_y' ($C_{y'_{max}}$) increases almost linearly in Z_s with remarkable consistency among the four experiments. The plotted line represents

$$C_{y'_{max}} = 0.48(1 + 2 Z_s/Z_i). \quad (1)$$

This means that for a source in the middle of the mixing layer, the maximum ground concentration is twice that given by unmodified Gaussian modeling. This should be a significant consideration to diffusion modelers.

5. Generalized urban effects on convective diffusion

Although there has been much study of the urban heat island and effects of cities on wind and turbulence, these physical effects are seldom introduced into diffusion modeling. Usually, urban diffusion measurements are analyzed using the conventional diffusion categorizing schemes discussed

at the beginning of this paper. However, in the framework of convective scaling, it is easy to anticipate the broader urban effects on diffusion, such as more rapid growth of σ_y and σ_z . As suggested by Weil and Brower (1982), convective scaling results imply that a good daytime stability index is w^*/U . Turbulence is predominantly convective when $|z_i/L| > 10$ or $w^*/u^* > 3$, which corresponds roughly to Pasquill "C" or more unstable conditions. In such cases, the spread of σ_y is proportional to w^*/U at small x , as is the descent distance of passive, elevated plumes, with the maximum ground concentrations near $x = 2 z_s/(w^*/U)$. In urban studies like that of McElroy and Pooler (1968), we see roughly linear growth in σ_z with x from a surface source before z_i is reached. For the most unstable classes, it is reasonable to assume that these σ_z are proportional to $(w^*/U)x$. How does a large urban area affect w^*/U ? There are three possibilities: U can be decreased, and both z_i and H^* can be increased. Below we consider each of these separately, then together.

The increased heat flux in an urban area has been well documented. Ching, et. al., (1983a), during their 1976 study of St. Louis, noted that average midday values of heat flux for the month of August were about 120 W/m^2 at a rural site, 135 W/m^2 in a residential area with many trees, and 295 W/m^2 at a commercial-industrial site nearly void of vegetation. At midday, the ratio of sensible heat flux to latent heat flux, the Bowen ratio, was about 2.0 at the industrial site and 0.6 at the rural site, showing the important role of vegetation in distributing solar energy. The Bowen ratio can be as small as 0.2 over wet cropland, with much-reduced sensible heat flux; the heat flux could be four times as large in

a relatively dry city. More typically, values of urban H^* are about twice rural values. The flux of latent heat also makes a small contribution to the buoyancy flux H^* , because water vapor is lighter than air. The ratio of this contribution compared to $(g/T)\overline{w'T'}$, the dry heat flux, is about 0.07 divided by the Bowen ratio. Because of the reduced effectiveness of latent heat in producing buoyancy, in heavily irrigated desert cities there may be a reversed urban effect; H^* may be less within the city, because a substantial fraction of insolation provides energy for evaporation of sprayed water and transpiration from vegetation.

The haze layer sometimes forms a visible dome over a city, suggesting an enhanced mixing depth. Wong and Dirks (1978) reported a July day in St. Louis with a measured z_i about 40% larger in the city than at the rural rawinsonde station. However, a systematic 30-month rawinsonde study of the same city showed much smaller rural-urban z_i differences (Schiermeier and Pooler, 1984). In fact, averages over all seasons showed no significant differences between the urban and the rural site's daytime z_i values. For the two summers studied, however, the average urban z_i at noon was about 13% larger than the rural z_i . During July-August expeditions with three rural rawinsonde sites, the average urban value of z_i at noon was 24% higher than the average rural value in 1975 and 5% higher in 1976. Thus, the average z_i enhancement at noon in summer in St. Louis varies greatly from summer to summer. Some of the lifting may be due to horizontal convergence as air entering a city slows down; it will be shown that this might be significant only for small U over a large city. It seems likely that the enhancement is mostly z_i growth caused by enhanced heating as air flows into an urban area. If all factors such as upper stability, over-city fetch (Δx), and heat flux ratios were fixed, the percentage increase

in z_i would be approximately proportional to $\Delta x/U$ (i.e., the time spent accumulating the enhanced heat flux). Again, small windspeeds over large cities would have the most effect on z_i .

Diffusion as a function of distance may also be increased by a decrease in wind speed; this may result from larger surface roughness in a city; surface drag is further increased by more vigorous vertical mixing of horizontal momentum caused by increased H^* . A typical urban value of roughness length, z_0 , is 1 m, although it may be several times as large in areas with tall buildings. This contrasts with the $z_0 \sim 0.3$ m typical of suburbs and agricultural land with scattered trees. The effects on wind speed of changes in z_0 and H^* can be conveniently explored using a surface layer similarity relationship for $\bar{u}/u^*$ based on an analysis of the Minnesota 1973 experimental data (Briggs, 1982a):

$$k \bar{u}/u^* = \ln(z/z_0'), \quad z_0' = z_0(1-4z/L)^{0.6}. \quad (2)$$

This relationship fits the wind profiles to $z \approx 0.2 z_i$; because there is little wind shear in the mixing layer above 100 m in convective conditions, we can apply (2) at a height $z = 100$ m, to approximate U with $\bar{u}$. Let us assume a certain initial speed U_0 , coming into the city over a roughness of 0.3 m, with $k = 0.4$ and $H^* = 25 \text{ cm}^2/\text{sec}^3$, a typical summer value over transpiring vegetation ($\bar{u}/u^*$ depends only weakly on H^* , so the value chosen is not critical). First, we can calculate the initial value of friction velocity, u_0^* , using (2). We assume that H^* doubles and z_0 becomes 1 m in the city (this could as well be the case for air encountering a large forest). The mixing layer has much inertia and U does not significantly decrease until long after the lower part of the wind profile adjusts to

the new surface. We can calculate a new value of u^* , u_1^* , using (2) with U_0 and the new surface parameters. Assuming that there is enough synoptic forcing to maintain the wind against a surface friction equal to u_0^{*2} , the initial frictional acceleration of the vertically-integrated mixing layer will be $z_i dU/dt = \Delta u^{*2} = (u_0^{*2} - u_1^{*2})$. Finally, we can use (2) once again to calculate the new asymptotic windspeed, U_∞ , that would occur if there were enough time or fetch to adjust back to $u^* = u_0^*$ over the new surface. Table 4 uses (2) to estimate these parameters, not with the precision of a planetary boundary-layer model, but with sufficient accuracy to estimate the magnitude of the urban effects.

The second-to-last line of Table 4 gives an idea of how long it takes U to adjust to a new value in equilibrium with u_0^* , changing at its initial rate. Since $|dU/dt|$ will diminish as U readjusts, it will actually take longer. It is clear that, on urban scales, there is not enough fetch for U to attain a new equilibrium value. Note that $(U_0/u_0^*)^2 z_i$, the ratio of upstream momentum flux in the mixing layer to upstream surface drag, gives a good estimate of the fetch required for adjustment. There is some significant deceleration across large cities; we find $dU/dx \approx -0.24$ m/sec per 10 km travel for all but the highest U . This is significant only for slow winds traversing large cities. Furthermore, if z_i is doubled, this change rate is halved. Notice that the change in z_0 causes about 3 times as much change in $U_0 - U_\infty$ as the change in H^* .

The net change in the diffusion rate, as characterized by w^*/U , obviously depends on U_0 and on city size (a fact that is ignored in most urban diffusion modeling). For a large city, with $U_0 = 2$ m/sec and a 20-km fetch, U may drop by 0.5 m/s and z_i may increase by 30%, because there is more time to

accumulate excess heat flux at low wind speeds (also, U_{zi} tends to maintain its initial value, to maintain the volume transport, except to the degree that the flow diverges horizontally around the city). Combining these changes with a doubling of the heat flux, the net factor of increase in w^*/U is $(1.3 \cdot 2)^{1/3} (2/1.5) = 1.38 (1.33) \approx 1.8$. We can expect a near doubling of σ_y and σ_z versus x compared with rural areas in these Pasquill "A-B" conditions. For this same fetch with $U_0 = 4$ m/s, z_i may increase by only 15% and U is again reduced by about 0.5 m/s, and the factor of increase in w^*/U is $(1.15 \cdot 2)^{1/3} (4/3.5) = 1.32 (1.14) \approx 1.5$. Diffusion enhancement is less in moderately convective conditions. At $U = 8$ m/s, mechanical turbulence becomes a significant factor, and the role of increased z_0 is more important.

Note from Table 4 that L increases only moderately as air encounters the city, due to compensating increases in u^* and H^* . L increases more at low wind speeds, but so does z_i , so the boundary-layer convective index $|z_i/L|$ does not change significantly. There are significant increases in both w^*/U and u^*/U , which are nondimensional indices of convective diffusion and mechanical diffusion rates, respectively. Yet they usually increase about equally, so that the character of diffusion is not changed much by the city, except in the layer disturbed by flow around buildings.

Although the daytime character of diffusion in a city may not differ much from that in rural areas, its nighttime character is far different, due mostly to the heat island. Could mixed-layer turbulence in a city ever be predominantly convective at night? The summertime heat flux measured in St. Louis was generally near zero in the city at night, but

on one evening it remained positive at about 40 W/m^2 at an industrial site until after midnight (Clarke et. al., 1982). Combining this with a typical nighttime mixing depth of 150 m in that city, $w^* = 0.44 \text{ m/s}$, about 1/4 of the daytime value. To consider the turbulence predominantly convective, we would require $|z_i/L| > 10$, or $w^* > 3u^*$, so u^* would have to be less than 0.15 m/s. Equation (2) then indicates that $u < 0.8 \text{ m/s}$ at 30 m. It seems unlikely that such low flow speeds would be associated with a strong positive urban H^* , because they may occur only during cold advection. Except in rare circumstances, it appears that nighttime turbulence in the urban mixed layer is primarily mechanical.

The effects discussed in this section are not limited to urban areas. Ching, et. al. (1983b) reported aircraft measurements over rural areas in which the heat flux was 2.0 to 3.5 times as large over a forested area as over irrigated cropland. The roughness length may be on the order of 1 m over forest and 0.1 m over cropland. If these surface inhomogeneities have scales on the same order as city inhomogeneities, then they should have similar effects on turbulence and diffusion.

6. Response of turbulence to heat flux inhomogeneities

The previous section estimated the changes in z_i and w^* on a broad urban scale produced by increased heat flux and roughness, and also estimated the rate of change in U after urban surfaces are encountered. Flights over St. Louis with turbulence-instrumented aircraft, reported by Ching et. al. (1984), showed that turbulent fluxes at 150 m are highly inhomogeneous. On one day, measured heat fluxes averaged over 20 km flight segments ranged from 20 MW over the Mississippi River to

6 times as large over the downwind side of the city. What kind of fetch over changed H^* is required for convective turbulence to respond to changed w^* ? The response distance for U is on the order of $(U_0/u_0^*)^2 z_i$, a large distance, compared with urban scales. It will be argued that the response distance for turbulence energy readjustments is on the order of several $(U_0/w_0^*) z_i$, a much shorter distance. Because this is the convective scaling distance, it so also indicates the distance of travel needed for complete vertical mixing. This applies to the mixing of horizontal momentum, too; thus, several $(U_0/w_0^*) z_i$ is also the distance it takes the wind profile to substantially adjust to a changed surface in convective conditions.

Suppose that a larger heat flux is encountered, so w^* is increased over its initial value, w_0^* . Turbulence velocities do not respond immediately, because the increased heat flux must rise all the way through z_i to release its potential energy, which is the driving force for the turbulence. From Willis and Deardorff (1976, Fig. 2), we see that the rise rate of the mean particle height, $\bar{z}$, is about $0.5 w^*$ for a surface release. Because heat flux arises from the surface, and the upward velocities do not respond to the new w^* immediately, it takes approximately $z_i/(0.5 w_0^*)$ in time for the new heat flux to rise through the whole layer. The response distance for the new heat flux to be fully effective is then about $2(U_0/w_0^*) z_i$.

What happens if heat flux is suddenly absent, as when air passes over a cool body of water, for instance? Turbulence energy in a steady convective boundary layer dissipates at a rate about $0.5 H^*$, balancing the mean rate of production of energy. This can also be expressed as $a\sigma^3/z_i$, where $\sigma^2 = \sigma_u^2 + \sigma_v^2 + \sigma_w^2$, the total variance. Each component of

σ averages approximately $0.6 w^*$ through the mixing layer, so $\sigma_0^2/w^{*2} = 3(0.6)^2$, $\sigma_0^3 = 1.12 H^* z_i$ and $a = 0.45$ in the case of steady H^* . When H^* is removed at $t = 0$, let us assume that $d(\sigma^2/2)/dt = -a\sigma^3/z_i$, where $\sigma^2/2$ is the total specific kinetic energy of turbulence. This equation has a simple solution:

$$\sigma/\sigma_0 = (1 + a\sigma_0 t/z_i)^{-1}. \quad (3)$$

Thus, the velocity variance reduces to half of its initial value at a time $t = z_i/(a\sigma_0) = 2.2 z_i/w_0^*$, or in a distance $x = 2 (U_0/w_0^*) z_i$. Thus, $X \sim 2$ seems to be generally the fetch required for substantial response to changes in heat flux.

How well do the velocity variances measured in the St. Louis program, referenced at the beginning of this section, respond to changes in w^* ? Any change in z_i over the city must be neglected because z_i was measured in only one location; z_i was about 1000 m on 18 Aug., 1976 and 1500 m the following day. Winds were from the E to NE. On both days, the aircraft measured local values of σ_u/w^* and of σ_w/w^* at $z = 150$ m were within $\pm 10\%$ of their mean values over approximately 80% of the area. The mean values of these ratios depend slightly on how heat flux measurements are extrapolated to the surface. If $(H_0 - H)/H_0$ is taken to be $2.4 z/z_i$ below $0.2 z_i$, an assumption supportable by Caughey and Palmer (1979, Fig. 3), then σ_u/w^* averaged 0.63 and 0.65 on the two days, and σ_w/w^* averaged 0.62 and 0.60. These are well within the ranges of other experimental values. The overflight values of σ_w/w^* were very close to 0.6 over areas of high heat flux in and around the city on both days. The ratio most deviated in areas of low heat flux, where σ_w did not drop as much as w^* . The most extreme example was on the 18th, on a N-S flight leg centered over

the Mississippi River at Mosentheim and Gaberet islands, just NE of the city center. These islands are heavily wooded and largely undeveloped, being mostly less than 5 m above the river, but both mainland shores are heavily developed. This island area, including water surface, is about 4 km wide and 12 km long, oriented NNE. The measured heat flux over this area was only 20 W/m^2 , compared with 60 W/m^2 measured upwind. The observed σ_w showed no drop at all, but remained within 5% of upwind and downwind values. The same was true on the 19th, but the heat flux varied only half as much (the patterns on the two days were quite similar). For the center of these islands, the fetch of the easterly wind from the developed shore was about 2 km. With $w_0^* \approx 1.35 \text{ m/s}$, $U_0 \approx 3 \text{ m/s}$, and $z_i = 1000 \text{ m}$, this gives $X \approx 0.9$. Apparently, this was not enough fetch to measurably affect σ_w at $0.15 z_i$ (some of the heat flux from the upwind shore would still be rising and driving turbulence at this point). The patterns for σ_u were similar to the σ_w patterns, reinforcing the above conclusion.

The best measurements of turbulence at a high heat flux site were of σ_v on a tower 1 km west of the Mississippi River. This commercial-industrial site was largely surrounded by low buildings and blacktop, with almost no vegetation. On eight days with easterly winds, the heat fluxes measured here were 3 to 4 times those measured at an upwind, rural tower. This probably was caused by rapid heat transfer to air cooled by an overwater trajectory. The rural value of σ_v/w^* averaged 0.62, agreeing with other experimental values. However, at the commercial-industrial site σ_v increased by about 15%, while w^* increased by about 50%; thus, σ_v/w^* averaged only 0.47 at this site. This partial response to increased w^* seems reasonable considering the modest dimensionless fetch from the Mississippi shoreline on these days, roughly $X \approx 1/2$.

7. The Organizing effect of heat flux inhomogeneities

An interesting phenomenon seldom discussed outside the context of shoreline flows is the ability of heat flux inhomogeneities sometimes to "organize" convective flow structures. A striking example of this was seen in the St. Louis overflights discussed in the previous section. The patterns of heat flux, H , and of the correlation coefficient for vertical velocity and temperature, R_{WT} , were very similar. This was especially well marked on the 18th, with its lower z_i , smaller U , and larger range of heat fluxes. R_{WT} reached a remarkable value of 0.8 on the downwind side of the city, almost coinciding with the heat flux peak of 120 W/m^2 . The contour at $R_{WT} = 0.7$ almost matched the contour at 100 W/m^2 . Upwind, over the islands where H was 20 MW, R_{WT} plummeted to 0.2, a very low correlation; values of 0.4 to 0.6 were typical of the rural areas. It may be that the convective eddies over the river, having lost their driving force, also lost their organization, becoming dissipative in nature. $R_{WT} = 0.8$ indicates a high degree of organization of vertical motions by heat, or vice versa. The latent heat flux, LE , and the correlation of vertical velocity with humidity fluctuations, R_{Wq} , also mimicked each other in their patterns, but these were almost reversed from the H and R_{WT} patterns. On the 18th, R_{Wq} reached a maximum at about 0.5 over the area of largest LE , 30 km upwind of the city. Here, the inverse Bowen ratio, LE/H , was approximately 5, and humidity fluctuations helped drive the turbulence. R_{Wq} fell to an extremely low value, 0.1, at precisely the point where R_{WT} was maximal. This was also the driest area, with $LE/H \approx 0.8$. It appears that dry heat organized and drove the convection in this area, while humidity took a passive role. Latent heat may have come mostly from

trees in this area, which were not the source of large heat fluxes. The areal distribution of H sources and LE sources may have also been poorly correlated within the city.

An even more graphic example of the organization of convection by heat flux inhomogeneities in St. Louis is shown in a paper by Kropfli and Kohn (1978). Metallicized chaff, just as was used in the BAO experiment, was released upwind of the city. It was observed by two doppler X-band radars. The analysis focused on the same horizontal planes, so that the radial velocities that each radar measured could be combined into a horizontal velocity vector. The mean velocity over the area was subtracted to get the eddy field, and the convergence field was calculated. This field was applied with the continuity equation integrated over z to infer vertical velocities. With this technique, Kropfli and Kohn discovered persistent, stationary horizontal rolls aligned with surface features of presumed heat flux anomalies. They showed the average patterns observed for two periods: 70 min on 28 July 1975, with $U = 4$ m/s from 20° and $z_i = 1.55$ km, and 20 min on 4 Aug., with $U = 4.7$ m/s from 42° and $z_i = 1.3$ km. Both days showed a line of convergence near the surface, with a corresponding updraft, originating in the SW area of Granite City (Fig. 5). This line maintained itself for the < 1 -km trajectory across the Mississippi and continued over the heavily commercial and industrial part of St. Louis. Auer (1978) classified the southern portion of Granite City as light to moderate industrial, with $< 5\%$ vegetation.

The case of 4 Aug., which had very little wind shear, showed another interesting feature. There was a persistent line of divergence and associated downdraft in the NW part of the city, as shown in Fig. 5. It was directly downwind of the Mosentheim and Gabaret islands area, in which very small heat fluxes were observed the following year. Furthermore, the wind was nearly aligned with the long axis of this area. It seems likely that, just as the Granite City area initiated upward motion, the river area of small H initiated the downdraft. The distance between these flow features, about 5 km, or $4 z_i$, is about four times that usually observed between adjacent updrafts and downdrafts in convective rolls (LeMone, 1973). Therefore, it seems likely that the flow features really were tied to surface features, especially since they did not move. The circle through the mean wind speed arrow in Fig. 5 shows the area of strongest downdraft, which was about 0.7 m/s at 600 m above the ground. A persistent downdraft of this magnitude can certainly be expected to intensify ground concentrations from elevated sources. During a diffusion experiment, I observed, from an airplane, a plume from a 900 MW power plant, emitted from a 250-m stack. It subsided directly over the waters of the Clinch River (TN), which is less than 1 km wide. This subsidence persisted for about an hour, on a very convective July day, when the wind direction happened to align with that section of the river.

I suggest that convective scales are relevant in these cases. It is not the surface temperature difference that directly drives such circulations (this difference can be 20 to 25°C, in the case of the Mississippi River at St. Louis). What drives the circulations is the temperature difference imparted to the ascending air, which is most directly dependent on heat

flux differences. A more direct physical argument is that release of potential energy drives the motion. This is proportional to the buoyancy flux difference, ΔH^* , times z_i and divided by the velocity scale; thus, $\Delta w^* = (\Delta H^* z_i)^{1/3}$ is the velocity scale for organized updrafts and downdrafts. For the day shown in Fig. 5, on the basis of the 1976 study, $\Delta H^* \sim 25 \text{ cm}^2/\text{s}^3$, so $w^* \approx 1.5 \text{ m/s}$. A typical magnitude of the measured downdraft was 0.6 m/s , which is $0.4 w^*$ in this case. This agrees with other estimates of downdraft magnitude (e.g., Briggs 1975) and is about the same magnitude as the convective turbulence downdrafts that cause higher ground concentrations from elevated sources than predicted by Gaussian models. However, the latter are intermittent and occur at a source point only about 60% of the time, whereas the rolls observed by Kropfli and Kohn (1978) were persistent, fixed features. The effect of a persistent downdraft on maximum ground concentrations from a point source might be modeled simply by subtracting $0.4(\Delta w^*/U)x$ from the effective source height, in addition to other model features.

Both fetch and wind alignment are important. It takes several nondimensional X , or $x = (U/w^*)z_i$, of travel over a heat flux anomaly for the full effect of Δw^* to take place. A small industrial area or lake may not initiate these flows except on low wind speed, convective days with relatively small z_i . Rivers, even the Mississippi, have little effect when a moderate or strong wind crosses them at a large angle. However, even a small river can cause persistent subsidence when the wind aligns with a straight section of the river, if its width is at least several tenths of z_i .

8. Summary and conclusions

This paper reviews the development of convective scaling and its success in ordering turbulence data from a wide variety of experiments. These included data collected by aircraft in field experiments using a tethered balloon in both the U.S.A. and England, in a laboratory tank experiment, and even in a numerical modeling experiment. Convective scaling was then applied to analysis of diffusion data, with consistent results from very dissimilar types of data bases, including a series of laboratory tank experiments and the Prairie Grass field experiment. Some of these results were surprising and not compatible with conventional diffusion modeling. The maximum concentrations from ground sources lifted off the ground after a travel distance of $0.5 (U/w^*)z_i$, and those from elevated sources had downward slopes of about $0.5 w^*/U$ until ground impact occurred. A few modelers used features observed in these experiments to develop more physically correct diffusion models, based on convective scaling. These models successfully predicted surface concentrations measured around several tall-stack sources with buoyant plumes.

Some preliminary results from a new field experiment are reported. The experiment was designed to test convective scaling in the atmosphere for diffusion from both elevated and surface passive tracers to distances of several $(U/w^*)z_i$. The plumes were mapped in three dimensions using two independent remote-sensing techniques: oil fog detected by lidar and metallicized chaff detected by X-band radar. The radar and chaff measurements from five averaging periods in 1982 have been analyzed. They were nondimensionalized using convective scaling and plotted in Figs. 1-4, along with surface-sampling data from elevated SF_6 releases from the

Netherlands and from the surface SO₂ releases of the Prairie Grass project, with mobile van and lidar measurements of the Morgantown power plant plume, and with curves derived from the laboratory tank and numerical modeling experiments. On the whole, the results were remarkably consistent, considering the diversity of experiments. One notable conclusion was that the maximum surface crosswind integrated concentration increased systematically with source height when compared with the standard Gaussian model prediction, $0.48Q/(Uz_s)$. The factor of increase was well approximated by $(1 + 2 z_s/z_i)$, so the surface concentration was double the Gaussian model prediction for a source in the middle of the mixed layer. This appeared to be caused by convective downdrafts.

The effects of heat flux and roughness length inhomogeneities on convective diffusion were considered. Using w^*/U as a convenient and physically direct index of convective diffusion rates, we explored its increase in urbanized areas due to enhanced heat fluxes, somewhat larger mixing depths, and reduced U , which is caused mostly by larger surface roughness. At low wind speeds in large cities, w^*/U approximately doubles. It changes less at larger wind speeds and in small cities. The convective character of the turbulence, as indicated by ratios like w^*/u^* , does not change much as air enters a city, even though the turbulence intensity is greater; w^* and u^* increase by approximately the same amount, except in areas of large roughness with little enhancement of heat flux (e.g., heavily vegetated areas). Although heat fluxes sometimes remain positive at night in very built-up areas, the turbulence in the nocturnal mixed layer is not predominantly convective except under extremely light winds, < 1 m/s. Large forests surrounded by cropland can have similar effects on turbulence and diffusion, such as enhancement of w^*/U .

Wind entering a city or a forest slows down but adjusts slowly, never reaching a new equilibrium speed, because the required fetch is on the order of $(U_0/u_0^*)^2 z_i$. In contrast to this, the fetch required for convective turbulence adjustments to changes in heat flux is characteristically $\sim 2 (U/w^*) z_i$. Turbulent variances measured at 150 m over the greater St. Louis area scaled to the local w^* within 10% of the expected values over about 80% of the area. The greatest departures were in areas of sharp heat flux contrasts, over and near the Mississippi River. The response of measured turbulence to the heat flux inhomogeneities may be crudely explained in terms of the dimensionless fetch, convectively scaled.

Finally, I discussed cases of highly organized convective motions near an urban heat flux maximum, as reflected in unusually large correlations between turbulent vertical velocities and temperature fluctuations; I also discussed cases of persistent and stationary convective rolls that appear to have been caused by heat flux anomalies. A line of surface convergence and updraft was probably initiated by an industrialized area, and a line of surface divergence and downdraft probably originated over a 4-km-wide area of river and low, wooded islands, surrounded by city. The persistent downdraft had velocities of about 0.6 m/s, enough to increase surface concentrations from elevated sources considerably. I suggest that convective scaling is relevant to these motions; the downdraft magnitude was about $0.4 \Delta w^*$. Such circulations are likely to occur near any river wider than 0.5 km on convective days if the wind direction is approximately aligned with the river.

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Tables

Table 1. Period parameters for September 1982 BAO experiment.

Table 2. Period parameters for convective days at Cabauw.

Table 3. Ranges of parameters in field experiments.

Table 4. Estimates of urban effects on U , u^* , and L .

Table 1. Period parameters for September 1982 BAO experiment

	Period (Date)				
	2(16)	3(16)	4(18)	5(20)	6(20)
Start time (MST)	13:04	14:11	13:54	11:53	13:12
Duration (min)	29	35	40	44	42
z_i (m)	520	730	960	980	1260
z_s (m)	235	235	167	3	3
z_s/z_i	0.45	0.32	0.174	0.003	0.003
H^* ($\text{cm}^2 \text{s}^{-3}$)	60	40	38	60	48
$w^* = (H^* z_i)^{1/3}$ (m s^{-1})	1.46	1.43	1.54	1.805	1.82
U (m s^{-1})	5.80	6.23	2.76	2.40	1.59
$z_i U/w^*$ (m)	2060	3180	1720	1300	1100
U/w^*	4.0	4.4	1.8	1.33	0.87
Duration $\div (z_i/U)$	19.4	17.9	6.9	6.5	3.2

Table 2. Period parameters for convective days at Cabauw

	Run				
	1a	1b	7a	9a	9b
z_i (m)	1200	1400	1450	1050	1050
z_s (m)	200	200	200	200	200
z_s/z_i	0.17	0.14	0.14	0.19	0.19
H^* ($\text{cm}^2 \text{s}^{-3}$)	19	11	43	46	38
$w^* = (H^* z_i)^{1/3}$ (m s^{-1})	1.33	1.14	1.84	1.69	1.59
U (m s^{-1})	7.68	7.62	7.80	6.48	6.08
$z_i U/w^*$ (m)	6950	9300	6150	4030	4030
U/w^*	5.8	6.7	4.2	3.8	3.8
Duration $\div (z_i/U)$	11.5	9.8	9.7	11.1	10.4

Table 3. Ranges of parameters in field experiments

	Experiment			
	BAO	Prairie Grass*	Cabauw	Morgantown
z_i (m)	520-1260	260-2100	1050-1450	1100-1650
U ($m s^{-1}$)	1.6-6.2	3.2-8.5	6.1-7.8	4.3-9.4
U/w^*	0.9-4.4	1.9-4.8	3.8-6.7	2.6-4.3
$X=(w^*U)x/z_i$	0.3-3.2	0.006-0.6	0.3-1.1	0.2-4.9
Duration $\div (z_i/U)$	3-19	1.7-9	10-12	14-25

* U measured at $z = 8$ m.Table 4. Estimates of urban effects on U , u^* , and L

	Assumed u_0 ($m s^{-1}$)			
	1	2	4	8
U_∞ ($m s^{-1}$)	0.43	1.1	2.6	5.7
$(U_0 - U_\infty)$ ($m s^{-1}$) due to:				
H^* change alone	-0.15	-0.22	-0.35	-0.45
z_0 change alone	-0.42	-0.66	-1.08	-1.85
$-L_0$ (m)	2.8	10.8	46	230
$-L_1$ (m)	4.6	14.6	55	250
u_0^* ($m s^{-1}$)	0.14	0.22	0.36	0.61
u_1^* ($m s^{-1}$)	0.21	0.31	0.48	0.79
$-\Delta u^{*2}$, m^2/s^2	0.024	0.046	0.100	0.25
For $z_i = 1000$ m, initial:				
$-dU/dt$ ($m s^{-1} h^{-1}$)	0.085	0.166	0.36	0.91
$-dU/dx$ ($m s^{-1}/10$ km)	0.24	0.23	0.25	0.32
$-(U_0 - U_\infty) \div dU/dx$ (km)	24	39	57	73
$0.45 (U_0/u_0^*)^2 z_i$ (km)	23	37	56	77

Figures

- Figure 1. Lateral dispersion, nondimensionalized by z_i , versus $(w^*/U)x/z_i$. "E" indicates elevated sources; other lines are near-surface releases. Flagged circles indicate Prairie Grass Run 1 (see text).
- Figure 2. Crosswind integrated concentration near the surface, nondimensionalized with z_i , U and Q , versus $(w^*/U)x/z_i$.
- Figure 3. Crosswind integrated concentration near the surface, nondimensionalized with z_s , U and Q versus $(w^*/U)x/z_s$. Numbers give z_s/z_i for each line or symbol.
- Figure 4. Maximum value of $C_y' = z_s U \int \chi dy / Q$ (left) and the distance of its occurrence (right), nondimensionalized by $(U/w^*)z_s$, as a function of nondimensional source height, $Z_s = z_s/z_i$.
- Figure 5. Eddy wind field at 0.3 km AGL in St. Louis on 4 Aug. 1975, as reported by Kropfli and Kohn (1978). The heavy arrow has been added to indicate mean wind velocity, and the open circle indicates the area of maximum downdraft velocity.

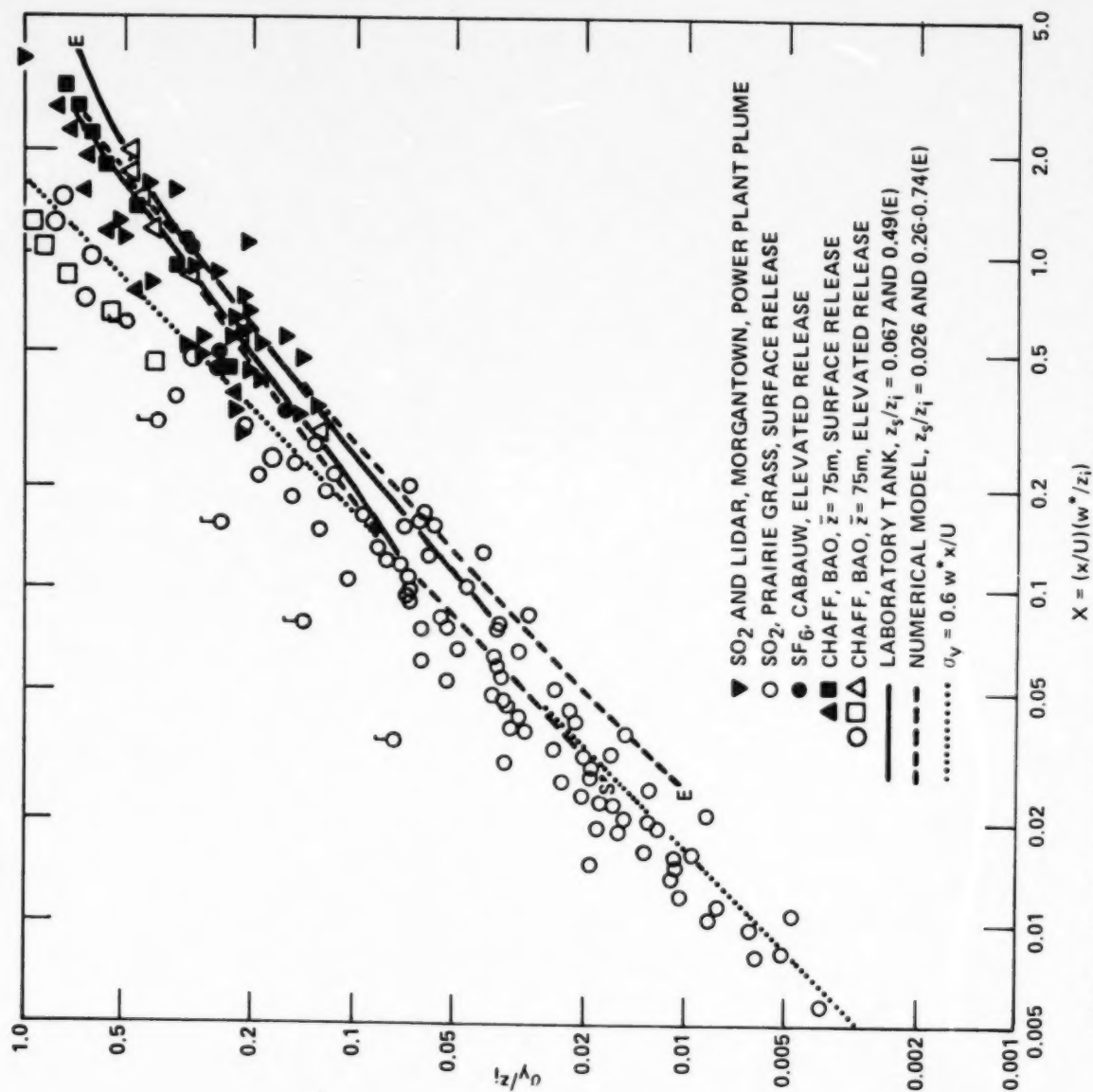


Figure 1. Lateral dispersion nondimensionalized by z_i versus $(w^*/U)x/z_i$. "E" indicates elevated sources; other lines are near-surface releases.

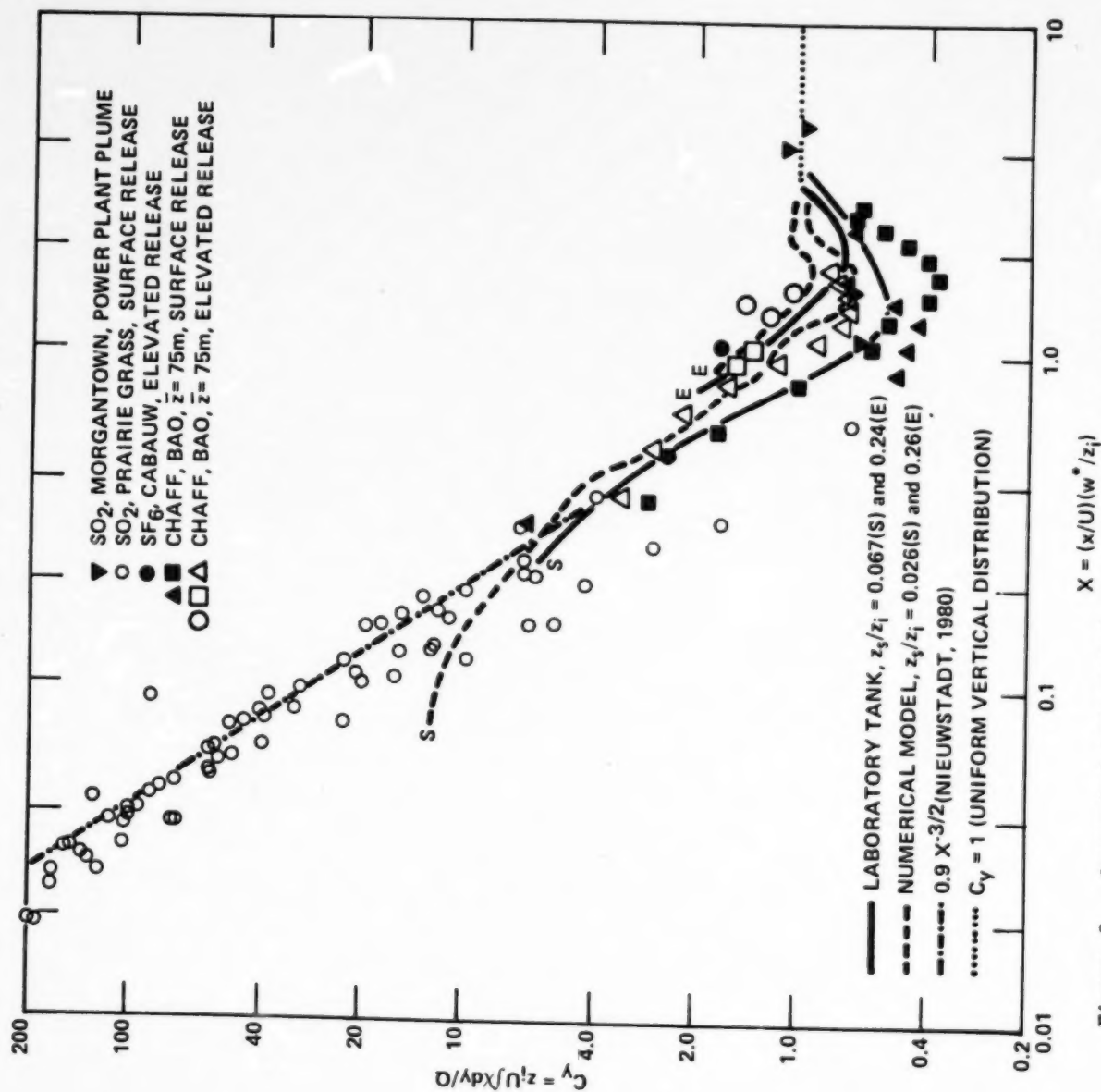


Figure 2. Crosswind integrated concentration near the surface nondimensionalized with z_1 , U and Q versus $(w^*/U)x/z_1$.

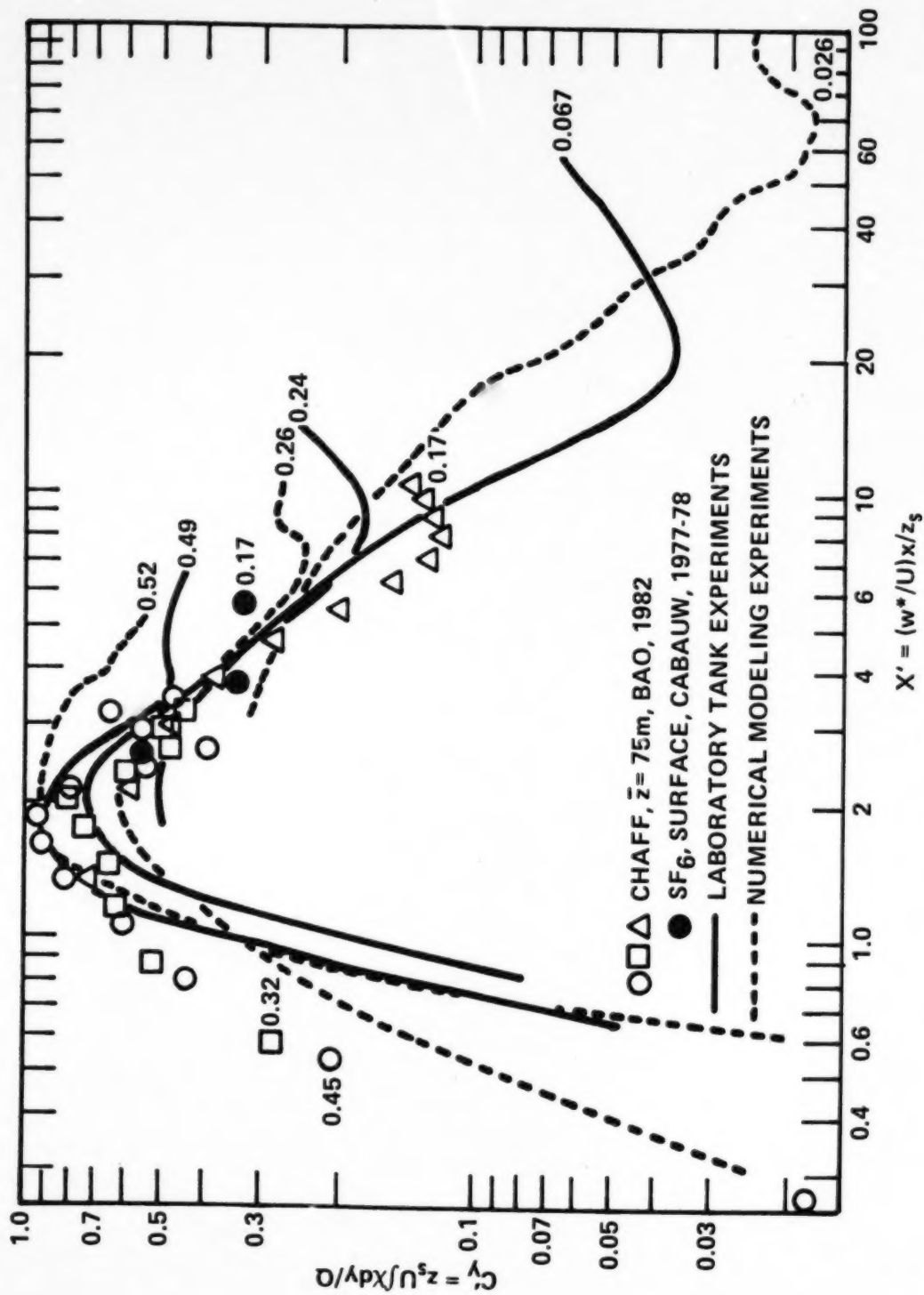


Figure 3. Crosswind integrated concentration near the surface nondimensionalized with z_s , U and Q versus $(w^*/U)x/z_s$.

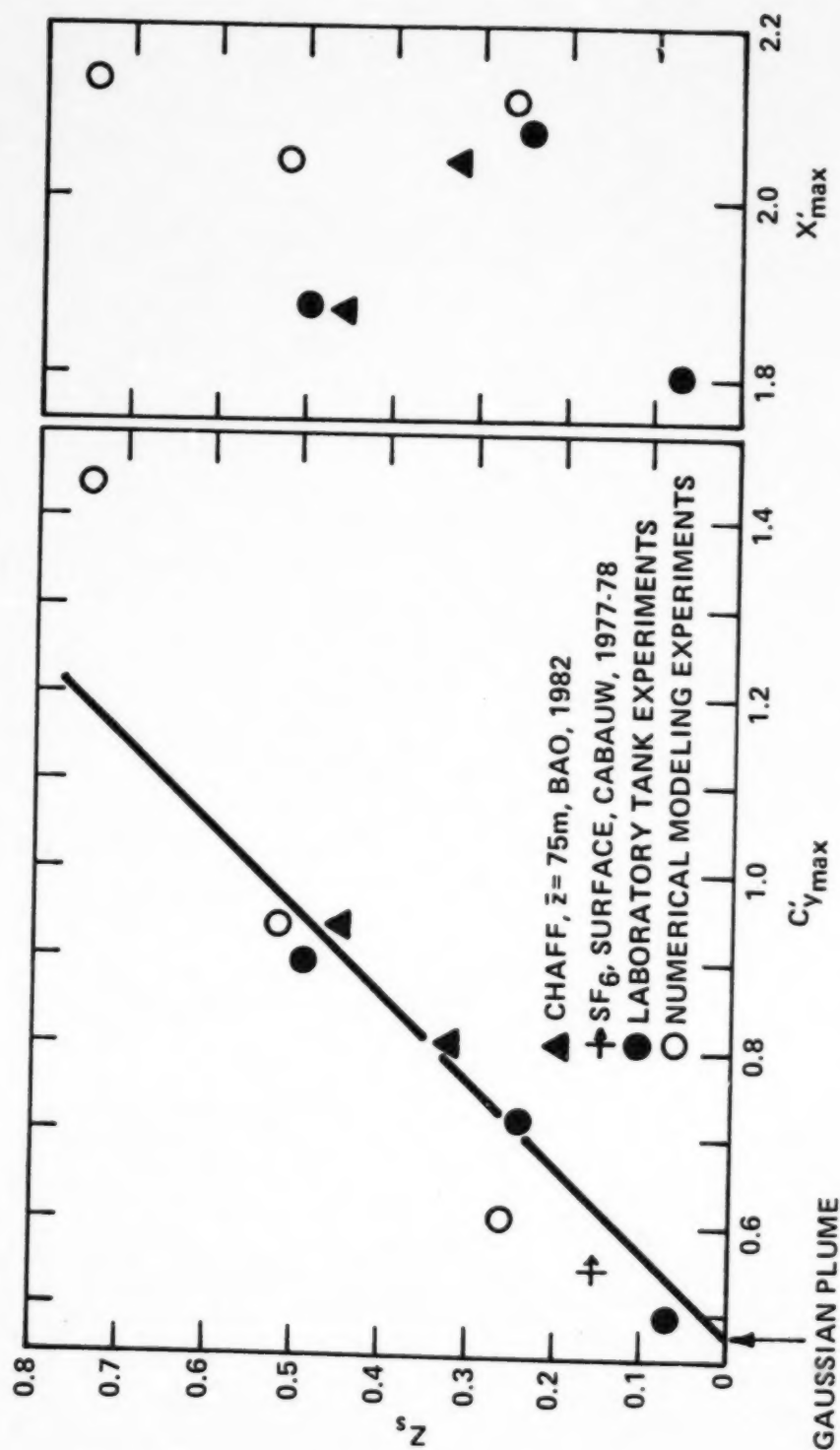


Figure 4. Maximum value of $C_y' = z_g U_f \chi_{dy} / Q$ and the distance of its occurrence, nondimensionalized by $(U/w^*)z_g$, as a function of nondimensional source height, $z_g = z_g/z_l$.

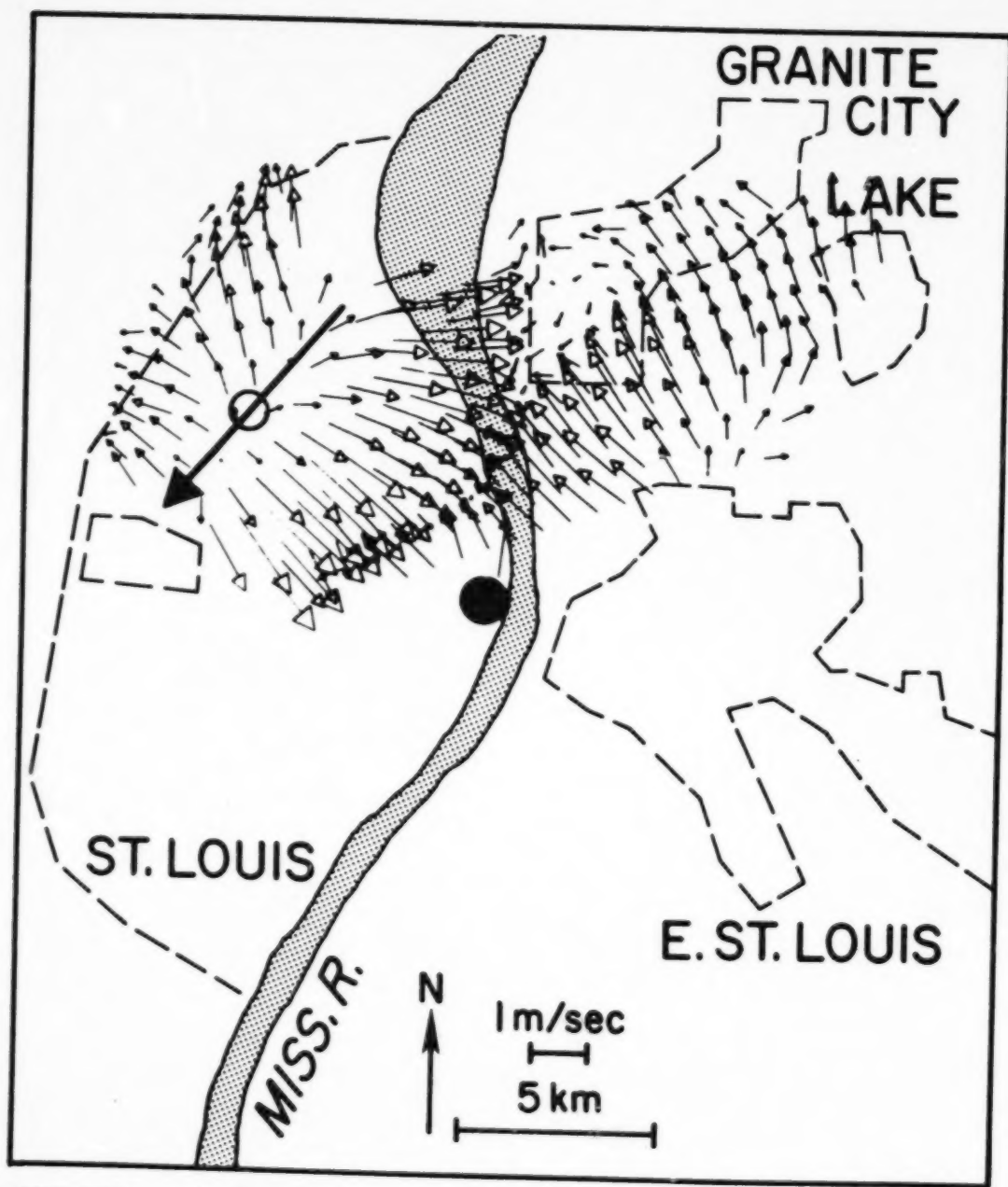


Figure 5. Eddy wind field at 0.3 km AGL in St. Louis on 4 August 1975, as reported by Kropfli and Kohn (1978). Heavy arrow has been added to indicate mean wind speed, and circle locates area of maximum downdraft velocity.